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The fragile climatological niche of the Baltic Sea

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The Baltic Sea is governed by the same physical laws as any other sea. Yet since it lies in a very different part of the temperature and salinity range than most other waters of the World Ocean, these laws result in dynamics that is not found elsewhere in the same extent. Here some consequences of this fact are discussed in light of recent observational and theoretical advances. We show that heat fluxes are almost negligible for restratification in this regime, and combine this idea with what is expected about the climatic warming to give a forecast of a substantial, non-linear change in the way the seasonal stratification and the vernal bloom in the Baltic Sea are formed. This discussion should also provide a conceptual framework for the discussion of the impacts of climate change on vernal bloom dynamics there, and in boreal estuaries in general.

Introduction

The existence of a density maximum in a saline water body results in dynamics that significantly differ from non-brackish parts of the oceans. The effect of heat forcing as well as heat gradients on density gradients is almost negligible in this regime. The free variation of temperature on isopycnals emerges as a null-space solution to the density-spice-space. The name of “shiver” is suggested for this null-space solution by Stipa (2002b).

The Baltic Sea is a major occupant of this very peculiar part of the oceanic temperature and salinity range. Specifically, in its present state it is the largest brackish water body that cools annually below the temperature of maximum density of sea water, θ_ρ (≈ 2–3 °C; cf. Fig. 1) (Voipio 1981, Mälkki and Tamsalu 1985). This results from the low, horizontally homogeneous salinity of the Baltic Sea as compared with many other estuaries; the salinity of its surface waters (above the permanent halocline at 60–80 m) ranges mostly between 5 and 10 psu, higher and lower values being concentrated to the Danish straits and the river mouths respectively. In addition, tidal stirring, which imposes a short time scale for stratification over many shelves, is virtually absent, helping to retain the signal of intrinsic physical processes in the observations. Our aim in this work is to present unpublished large-scale observational physical and biological evidence for the negligible effect of temperature on the surface layer dynamics in the spring restratification phase. We start by briefly reviewing the physical thermodynamics governing the cold

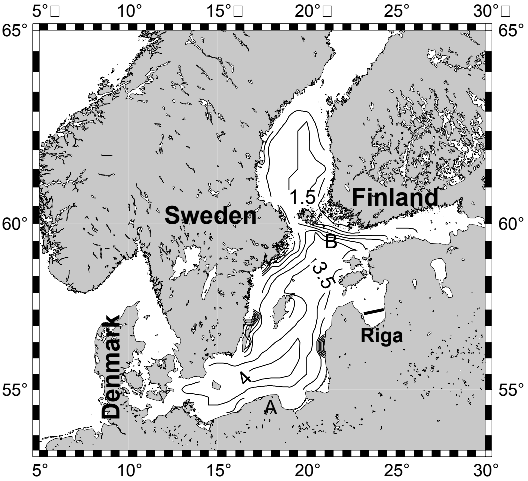

Fig. 1. Map of the Baltic Sea, with maximum sea surface temperature (°C) (Grönvall *et al.* 1987) on March 1 in the period 1965–1986. The mean and minimum temperatures are approximately 2 and 4 °C lower, respectively. A and B denote the end points of the transect shown in Fig. 3.

Baltic Sea and conclude by emphasizing that the Baltic Sea may be affected by a warmer climate more than a linear argument would indicate, both physically and ecologically. For a more thorough theoretical discussion, *see* Stipa (2002b) and Stipa (2002a).

Isopycnal gradients of θ and s

For hydrostatic, baroclinic, incompressible and mechanically unforced processes, density gradients drive the motion. The relative importance of temperature θ and salinity s on the dynamics is then a result of their effect on the density gradients:

$$\rho^{-1}\nabla\rho = \rho^{-1}\frac{\partial\rho}{\partial\theta}\nabla\theta + \rho^{-1}\frac{\partial\rho}{\partial s}\nabla s \equiv -\alpha\nabla\theta + \beta\nabla s,$$

where α is the thermal expansion and β the haline contraction coefficient. Along isopycnals, density gradients are by definition zero: $\rho^{-1}\nabla\rho \cdot \boldsymbol{e}_\rho \equiv 0$, where $\boldsymbol{e}_\rho$ is a vector on the tangent isopycnal plane.

However, density contains only a part of the information in the (θ, s) fields. Veronis (1972) and Mamayev (1975) introduced and Jackett and McDougall (1985) refined the variable for the remaining information as $d\tau = +\alpha d\theta + \beta ds$, frequently called "spice" after Munk (1981). Now we have a complete set of equations for isopycnal variations of temperature and salinity,

$$\rho^{-1}\nabla\rho\cdot\boldsymbol{e}_\rho = -\alpha\nabla\theta\cdot\boldsymbol{e}_\rho + \beta\nabla s\cdot\boldsymbol{e}_\rho \equiv 0 \qquad (1)$$

$$\alpha\nabla\theta\cdot\boldsymbol{e}_\rho + \beta\nabla s\cdot\boldsymbol{e}_\rho = \nabla\tau_\rho\cdot\boldsymbol{e}_\rho \qquad (2)$$

Stipa (2002b) noted the existence of a "null space" in the above equations when α passes through zero at θ_ρ; $\alpha(\theta_\rho) = 0$. Then, Eq. 1 gives $\beta\nabla s \cdot \boldsymbol{e}_\rho = 0$, which in turn renders the right-hand side in Eq. 2 zero. Isopycnals, and therefore the dynamics, are completely determined by salinity gradients in this range. Stipa (2002b) coined the name "shiver" for the free isopycnal temperature variance, since its importance increases in cold oceanic waters (e.g. the Arctic and Antarctic waters). Consequently, the choice of salt as the estuarine, fluid dynamical coordinate (Walin 1977) becomes a choice of an isopycnal coordinate system in the temperature range near θ_ρ.

Apparently because of the vanishing spice gradients, double-diffusive convection is not readily observed in the Baltic (e.g. Prandke and Stips 1992) as its effectiveness also depends on a non-zero α, as pointed out by Zhurbas and Paka (1999). This property may further make horizontal and vertical mixing processes in the Baltic simpler to analyze than in the ocean.

Restratification

Stratification in an unstratified sea is created by forcing through the surface or internal redistribution of mass. A simplified equation for the buoyancy tendency on a parcel in the surface layer can be written for a non-material control volume as

$$\frac{\tilde{D}b}{\tilde{D}t} = \frac{\alpha g}{C_p}\nabla\cdot\boldsymbol{J}_q + \beta s g\nabla\cdot\boldsymbol{J}_w - \nabla\cdot\overline{v'b'} \qquad (3)$$

where g is the gravitational acceleration, $\boldsymbol{J}_q$ the heat flux, $\boldsymbol{J}_w$ the freshwater flux, C_p the heat capacity at constant pressure and v' the velocity fluctuation relative to the control volume. $\frac{\tilde{D}}{\tilde{D}t}$ denotes the advective operator for the non-material control volume, cf. Stipa (2002a).

It is evident that restratification, i.e., creation of stratification, can take place by several mechanisms: (1) due to external fluxes caused by (a) vertical heating divergence or (b) dilution of salinity; or (2) internal redistribution of density surfaces via the divergence of eddy fluxes, $\nabla \cdot \overline{v'b'}$. Mechanism 1a is dominant for most parts of oceans (e.g. Carmack 2000) as well as freshwater lakes, and 2 has been shown to be important after deep oceanic convection by Jones and Marshall (1997), whereas a combination of 1a–b and 2 is relevant for estuaries (Hill 1998). 1b by itself has certain relevance at least for the tropical ocean (Brainerd and Gregg, 1997).

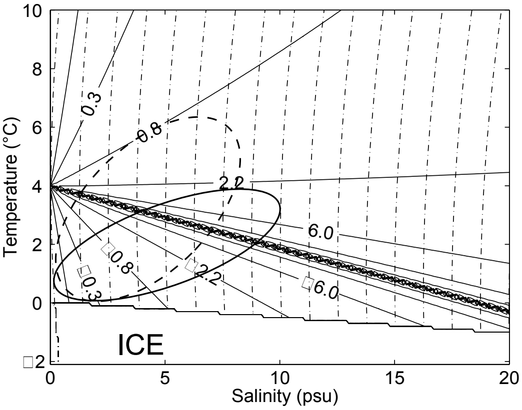

Fig. 2. The restratification parameter R_B as a function of temperature and salinity, scaled with F_0 = 10 kg m^{-2} d^{-1}, Q_0 = 50 W m^{-2}. The dash-dotted lines are contours of constant density. The numerical value indicates how much larger the contribution of the freshwater flux to the buoyancy forcing is with these scales in different temperatures and salinities, as compared to the contribution from the heat flux. The parameter goes through infinity at θ_ρ, showing the vanishing of the heat flux contribution. The solid ellipse represents a schematic distribution of surface waters in March in the present climate (Grönvall *et al.* 1987), and the dashed curve a projection for the 2 × CO$_2$ climate of IPCC (2001).

For the direct restratification in cold brackish waters, heat is unimportant and freshwater becomes dominant. An enlightening illustration of the importance of this effect can be found by considering the relative contribution of heating vs. salinity dilution in changing the surface density, which can be compactly described with the dimensionless restratification parameter

$$R_B \equiv \frac{\beta C_p S F_0}{\alpha Q_0} \tag{4}$$

F_0 and Q_0 being typical scales of freshwater and heat fluxes in the restratification phase. The parameter is smaller than one when heating dominates the density change, hence providing a way to discriminate between the strength of the competing effects in creating a stratified surface layer. Note that the large heat capacity of water ties rapid, small-amplitude heat input fluctuations to a large extent into internal heat, rendering their effect on the water temperature small. Figure 2 shows that at θ_ρ, $|R_B|$ goes through infinity and is large in a finite interval around it (approximately between temperatures 1 and 5 °C in the Baltic Sea), i.e., an infinite heat flux is needed to compensate for the smallest freshwater flux in this regime. This in turn means that a systematic addition of heat has a negligible effect on the density evolution — but not on temperature evolution — of surface waters in these conditions. Since the Baltic Sea cools in the winter to temperatures close to the freezing point, this also means that the mechanism by which the Baltic Sea is restratified is very different from the horizontally homogeneous heating of the extra-polar oceans and freshwater lakes. Since the dominant source of freshwater in the Baltic Proper is land runoff (Voipio 1981) (some originates from melting ice in the northern bays), the buoyancy supply during restratification is at the coasts, eventually spreading over the Baltic Proper. Knowledge of the exact mechanism by which the restratification takes place is crucial for the understanding of marine ecosystems and possibly our climate, since the main mechanism for drawdown of atmospheric carbon in high latitudes, the vernal bloom of phytoplankton (Lignell *et al.* 1993), is greatly dependent on the onset of stratification (Sverdrup 1953, Smetacek and Passow 1990).

There is in fact a great conceptual similarity between the Baltic restratification and the restratification that takes place after deep oceanic convection; there a low buoyancy anomaly is also surrounded by a high buoyancy reservoir. The restratification of such a buoyancy anomaly has recently been shown to be well described in terms of baroclinic instability of the buoyancy gradient (Jones and Marshall 1997). The difference is that in the Baltic Sea the buoyancy supply does not come from the semi-infinite stratified ocean around the homogenized area, but ultimately from freshwater sources around the coasts. Further, after oceanic convection the actual heat flux diminishes, whereas in the Baltic Sea α vanishes in addition, resulting in a short-term decoupling of the buoyancy forcing and the heat flux.

Large-scale observational evidence for the freshwater-driven restratification

For some time there has been observational evidence that restratification is indeed caused by lateral eddy-induced freshwater fluxes in the Baltic Sea (Kahru and Nõmmann 1990, Stipa *et al.* 1999); remarkably, the vernal bloom seems to be the most sensitive indicator of restratified areas.

A further proof for the importance of the eddy mechanism is found in the 1-dimensional mixed-layer model experiments of Eilola (1997) and Stigebrandt (1985), which do not reproduce the formation of "Baltic winter water" (BWW), a cold ($\theta < \theta_\rho$) layer capped beneath the seasonal pycnocline, correctly by considering only thermal stratification, without introducing a pulsating freshwater source. During the study period of Eilola (1997), there was no rain and almost no wind, so a lateral eddy flux is the most obvious explanation. It seems justifiable to propose that the very existence of the winter water layer, a prominent feature of Baltic Sea hydrography that cannot be capped by merely heating the surface, is a manifestation of the eddy-driven restratification mechanism.

Physical evidence

When a recently melted ($\theta \approx 0$ °C), boreal (dimictic) freshwater lake is heated by solar radiation from the surface, the whole water column is homogenized to θ_ρ, before the restratification starts at the surface. Hence the lowest temperatures found after winter in a lake are slightly but significantly above θ_ρ.

The Baltic Sea differs from lakes in this respect. The lowest temperatures found beneath the seasonal stratification are in many years significantly below θ_ρ. Figure 3 shows a south–north cross section of the Baltic Proper. A layer of water colder than 2 °C is found at intermediate depth, with the coldest temperature in the northern part of the basin. This water mass ("Baltic Winter Water") could not exist without the stratifying effect of freshwater eddies, which stratify the water column and inhibit the convection that would otherwise take place until the water warms to θ_ρ. Numerical evidence for this primarily horizontal process is given by Stipa (2002c); cf. Stipa (2002a). The disappearance of the BWW region towards the south may indicate the lesser importance of the freshwater-induced restratification mechanism in the southern Baltic. In addition to coastal river discharge, the freshwater released from the thicker ice cover in the northern Baltic may make a significant contribution to the freshwater budget; however, Eilola and Stigebrandt (1998) note the gradual southward progression of freshwater thickness during the course of the ice-free season. This progression is not explicable by e.g. Ekman transport, therefore, internal oceanic processes are likely the determining factors of southward freshwater progression. Because of the temperature and salinity properties of the layer in Figure 3, northward transport of BWW is not a likely explanation for the larger BWW thickness in the northern Baltic.

Biological evidence by remote sensing

As intriguing as this physical process is by itself, its greatest implications may be found in the preconditioning for the vernal phytoplankton bloom. There is ample evidence that a sufficient residence time in the well-lit surface layer is needed for the bloom commencement (Smetacek and Passow 1990), and the contradicting results known from the Baltic Sea (cf. Smetacek and Passow 1990) have only considered the temperature gradient as a stabilizing mechanism — which by the above discussion leads to misinterpretations of the initiation of the vernal bloom.

In combination with the small Rossby radius (Fennel *et al.* 1991) (3 km), the eddy nature of this process makes the stratification conditions in the Baltic Sea very patchy in spring, and in fact

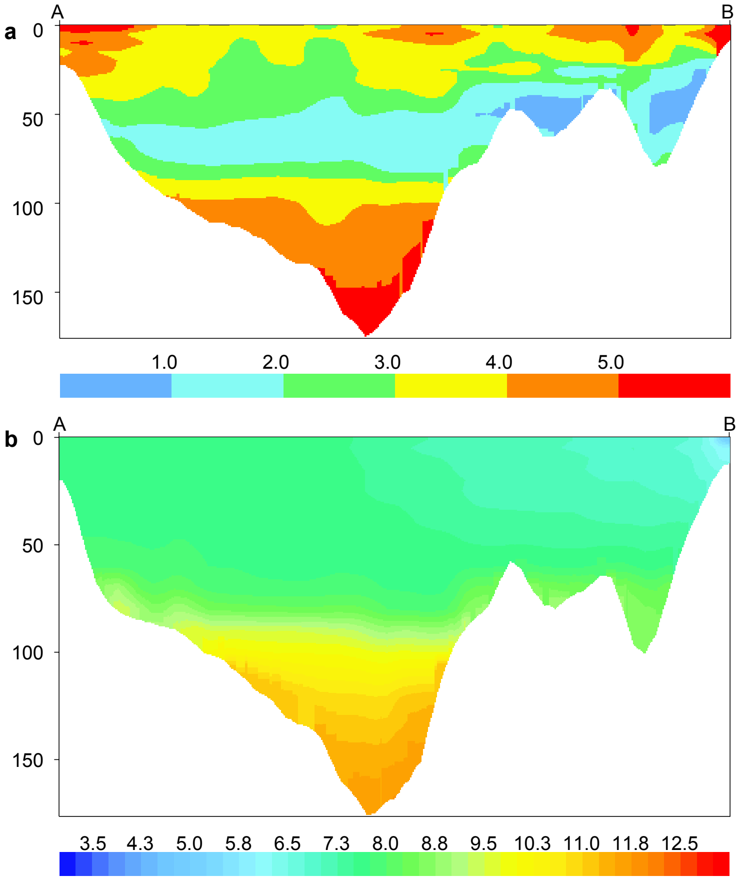

Fig. 3. A south (A)–north (B) transect of (**a**) temperature (C) and (**b**) salt across the Baltic Proper, cf. Fig. 1. Data from the Baltic Environmental Database between 1 April and 1 July 1987 (Sokolov *et al.* 1997). The winter water layer with $\theta < 2$ °C $< \theta_\rho$ is seen at 50–80 meters, with the lowest θ and largest thickness in the northern parts which have a stronger influence of freshwater.

give it a very characteristic and unique mosaic of vernal bloom (Kahru and Nõmmann 1990), that may not be encountered in the same spatial extent in other parts of the world. The physical patchiness is mostly visible in the salinity field, but unfortunately there are no operational methods yet to map the sea surface salinity. However, the theory suggests a decoupling between surface chlorophyll and temperature fields, which should be detectable by remote sensing.

The pigment field of the spring bloom as seen by satellite sensors (Fig. 4) indeed carries suggestive evidence for this process (*see* also Horstmann *et al.* (1986; Figs. 5 and 6)). The temperature in the figure is close to θ_ρ, i.e. the temperature has no effect on density, and in the absence of freshwater influence, the whole water column would be in a state of vigorous mixing and convection, as observed in lakes in this stage of the season. The restratification by freshwater eddies, driven by e.g. instability of salinity fronts, becomes visible because of the indirect effects on the vernal bloom via the generation of stratification. Witness the presence of large chlorophyll values in the northern Gulf of Riga, west of the southern tip of Finland and across the easternmost Gulf of Finland in Fig. 4. At least on the northern shore of the Gulf of Finland, there are considerable chlorophyll concentrations despite the fact that $\theta \approx \theta_\rho$. Note that the euphotic zone is less than ten meters, i.e. significantly shallower than the bottom depth of 50 meters, on the average. A cloud band, whose boundaries affect the chlorophyll algorithm, is observed in the middle of the Gulf of Finland.

Sensitivity to climate change

Yet, as can be seen from the parameter R_B in Fig. 2, this unique dynamical regime exists only due to relatively low salinity in combination with

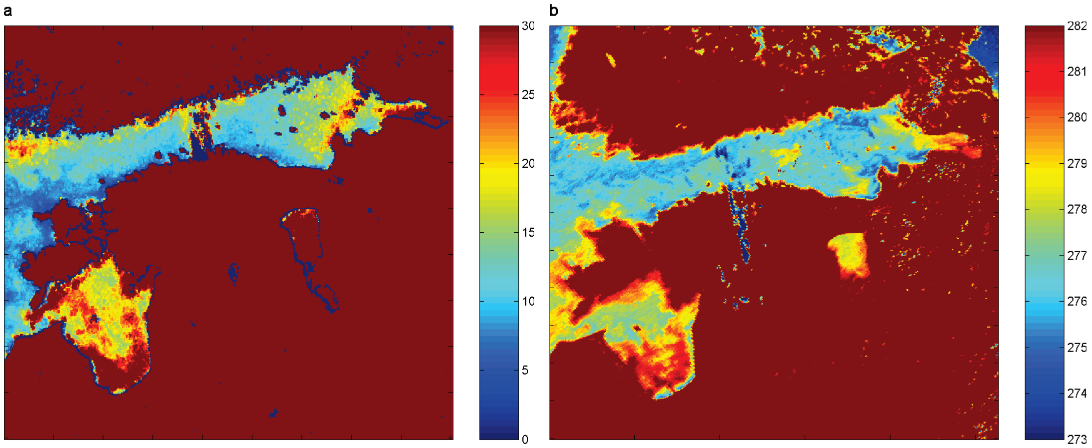

Fig. 4. A chlorophyll (μg l^{-1}) map (**a**) and a sea surface temperature (SST, K) map (**b**) from the Baltic Sea (Gulf of Finland). The chlorophyll and the SST maps are calculated from SeaWiFS and NOAA/AVHRR data, respectively. Both data are from satellite overpasses from 1 May 1999 just before 12 UTC. The amount of chlorophyll has been determined using channel ratio algorithm that is calibrated with operational flow-through fluorometer measurements (Leppanen and Rantajarvi 1995). The chlorophyll retrieval algorithm is available from J. Vepsäläinen. SST map is based on a split window method which compensates the effects of atmosphere (Pyhalahti 1998). In both **a** and **b**, dark blue areas are cloudy and red areas are land.

the minimum winter temperatures of the Baltic waters below θ_ρ, and it is highly sensitive to small changes in spring θ and s surface values but not to instantaneous heat or freshwater fluxes.

An analysis of historical data by Eilola (1997) shows that an increase of 2 °C in mean winter air temperature is able to bring the BWW temperatures above θ_ρ. The estimates of the increase in winter temperatures in the Baltic Sea region for a 2 × CO_2 climate (IPCC 2001) vary mostly between 3 and 5 °C. Thus, at least the southern parts of the Baltic can be expected to be shifted towards the oceanic, heat input dominated, horizontally more homogeneous restratification mode; in other words, the Baltic Sea will shiver less and even develop some spiciness. With the nonlinear restratification dynamics just described, this is a straightforward consequence of climatic warming and, because e.g. the vernal bloom critically depends on stratification conditions, it may have a profound impact on the biogeochemical cycles.

However, some parts of the Baltic Sea are expected to freeze in future years also (Meier 2001), so one may actually expect a more pronounced division of the Baltic Sea to a mostly thermally restratified, warm and spicy southern part and a freshwater restratified, cold and spiceless northern part. A schematic distribution of springtime Baltic surface waters in the temperature, salinity space is overlaid on Fig. 2, both for the present climate of Grönvall *et al.* (1987) and for the conditions anticipated for the scenario of Meier (2001).

As a result of the increasing importance of thermal restratification with rising SSTs, the role of eddies in the restratification process can be expected to decrease as well. The eddies transport coastal water offshore carrying tracers of coastal origin (Kahru and Nõmmann 1990), including nutrients for phytoplankton growth. Since this offshore flux is a function of the buoyancy gradient (Visbeck *et al.* 1997, Jones and Marshall 1997), it can be expected to change if the forcing by freshwater discharge changes. Recent studies (Oschlies and Garçon 1998, McGillicuddy *et al.* 1998) point to the importance of the eddies themselves on the ecosystem; future studies may be able to tell what the impact of a possibly decreasing eddy activity would be to the vernal bloom. It is also important to note, that this lateral restratification process implies the necessity to use fully 3-dimensional models to study the development of stratification in the Baltic Sea.

In conclusion, we have pointed to one aspect which makes the dynamics in the Baltic Sea different from almost any other water body in the world, and shown that this dynamical regime exists in a very narrow climatological niche currently occupied by the Baltic. While the consequences of climatic warming to the findings presented here may be difficult to alleviate, there is a need for immediate action at least on one front: future discussions of restratification and vernal bloom dynamics in the Baltic Sea should not be based on the thermal stratification only, and many existing discussions would deserve a closer inspection before climatological conclusions are drawn.

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